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SOME ASPECTS OF THE CIRCULATION OF MARS

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ABSTRACT

Estimates of the vertical temperature structure and heat balance of Mars are reviewed and compared with the corresponding quantities on the Earth. The probable resulting circulation is discussed, and reasons for expecting a stronger solstice circulation on Mars than on the Earth are given. The problem of thermally driven tides is reviewed. The amplitude of such tides is likely to be small.

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1. Introduction

The problem of the circulation of the Martian atmosphere is of great interest not only because of its close connection with a variety of other Martian problems, but also because of its implications for terrestrial meteorology. Both theory (Pedlosky, 1964; Phillips, 1963) and laboratory experiments on differentially heated rotating fluids (Fultz, 1961) indicate that the two important parameters determining dynamical similarity of the gross features of the flow are the Rossby number, $R_o = \frac{V}{fL}$, where V is a characteristic relative fluid velocity, f the Coriolis parameter, and L a characteristic scale of motion, and a static stability parameter, $\varkappa = \frac{\delta s}{c_p}$, with δs the characteristic change in specific entropy over the depth of the circulation system, and c_p the constant pressure specific heat. The horizontal gradient in heating, primarily, determines V; L is either the planetary radius or a somewhat smaller scale associated with possible instabilities of the flow. For stability of the atmosphere with respect to small scale vertical convection, $\varkappa$ must be positive; it is determined jointly by the heating field and by the motion in a complex way: convective heat input near the ground and radiative cooling in the upper atmosphere tend to diminish $\varkappa$; upward heat transport associated with large scale motions increases $\varkappa$. The resulting value is determined by a balance between these processes, but for a given field of heating by radiation and small scale convection, $\varkappa$ increases as the intensity of the large scale circulation increases; hence it is ultimately related to the horizontal heating gradient. (For a discussion of the possible relationship between $\varkappa$ and the gross circulations of both Mars and the

Earth as well as other aspects of the Mars problem, see Mintz, 1961.)

The Earth and Mars have nearly equal rotation rates, and nearly equal axial tilts. Since the axial tilt is the most important factor in determining the differential heating, we can expect similarities in the differential heating on the two planets. Furthermore, since the lower atmospheres of both planets are heated by small scale convection near the ground and lose heat by radiation at higher levels, the relationship between κ and the circulation intensity should be similar. In some respects, the problem of the general circulation of Mars may be much more straightforward than that of the Earth. Water vapor, oceans, and clouds, all of which tremendously complicate the terrestrial problem, need not be considered on Mars. Consequently, the relationship between the external parameters: solar heat input, rotation rate, tilt, etc., and the resulting circulation should be simpler on Mars.

2. The vertical temperature distribution

Since differential heating is an essential ingredient in the general circulation recipe, it will be necessary to consider its magnitude and distribution. It is helpful, however, to first review some of the main features of the expected vertical distribution of temperature as given by radiative equilibrium calculations. The temperature distribution derived from the radiative equilibrium hypothesis ought to give at least a rough qualitative picture of the real distribution. Furthermore, the calculations give some insight into the role played by radiation and small scale convection in determining the vertical temperature distribution, even though the large scale motions would modify this structure.

Figure 1 shows a combination of separate calculations by Goody (1957) for the lower atmosphere and by Chamberlain (1962) for the upper atmosphere, as compared with the Earth. Several significant differences are evident: the Martian tropopause is higher, and the stratosphere is significantly colder. The temperature peak at the terrestrial 50 km level, which occurs as a result of absorption of solar radiation by ozone, does not appear on Mars. As we shall see, this fact may be significant for the problem of atmospheric tides. Chamberlain's calculations predict a deep temperature minimum near 130 km corresponding to the breakdown of Kirchhoff's Law for CO_2 below this height. In this region, convective equilibrium should determine the temperature distribution resulting in the relatively steep dry adiabatic lapse rate. Chamberlain concludes that the minimum temperature may be low enough for CO_2 to condense.

Despite these differences, the lower atmospheres of both planets are characterized by temperatures that decrease sharply with height. This is a consequence of the fact that most of the incoming solar energy in both cases is absorbed at the surface, rather than within the atmosphere. Radiation alone would produce a temperature discontinuity at the ground. It is assumed that small scale convection would smooth out the discontinuity and lead to an adiabatic lapse rate in the lower troposphere.

More recent detailed calculations for the lower atmosphere have been made by Prabhakara and Hogan (1965). They took into account the absorption of solar radiation as well as infrared emission by all of the important carbon dioxide bands, and also considered the possibility

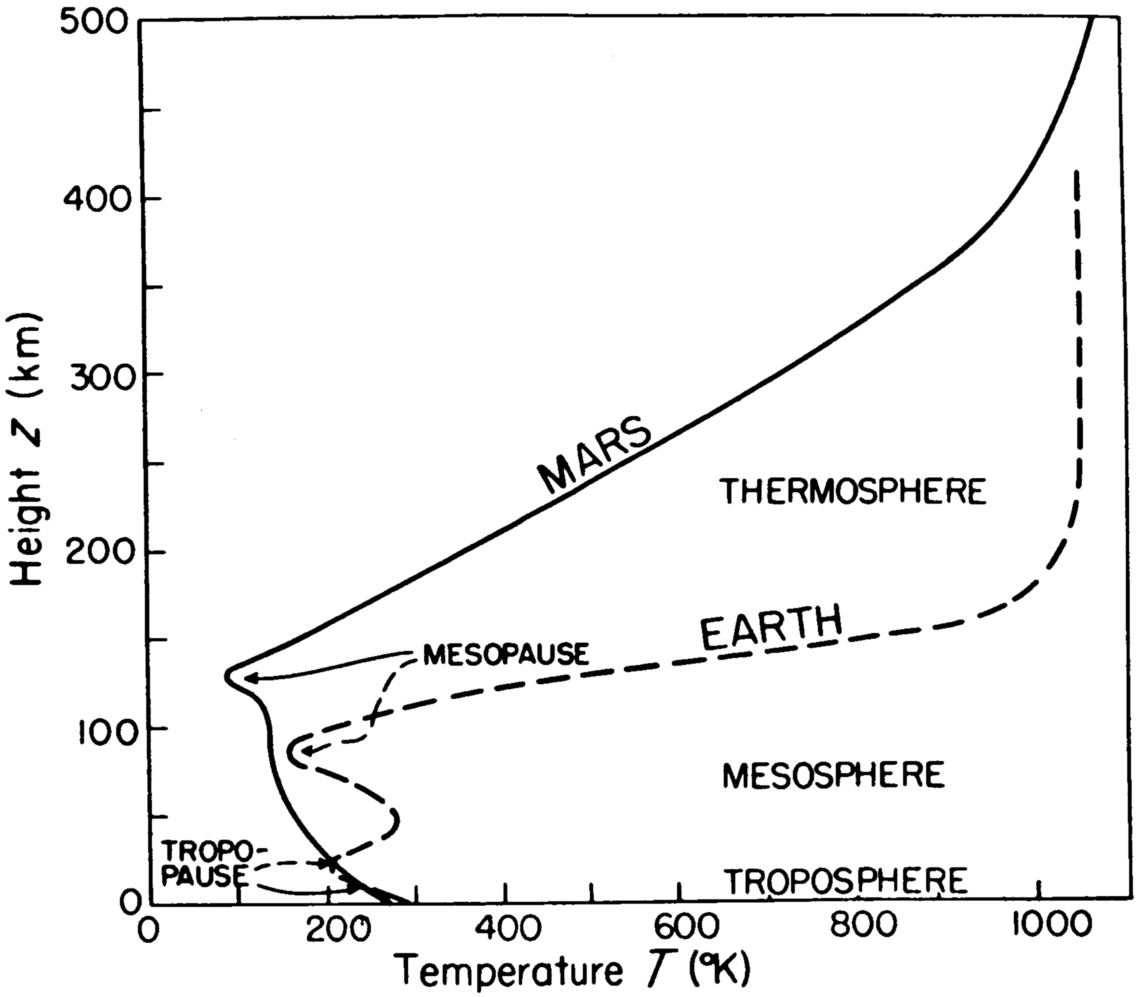
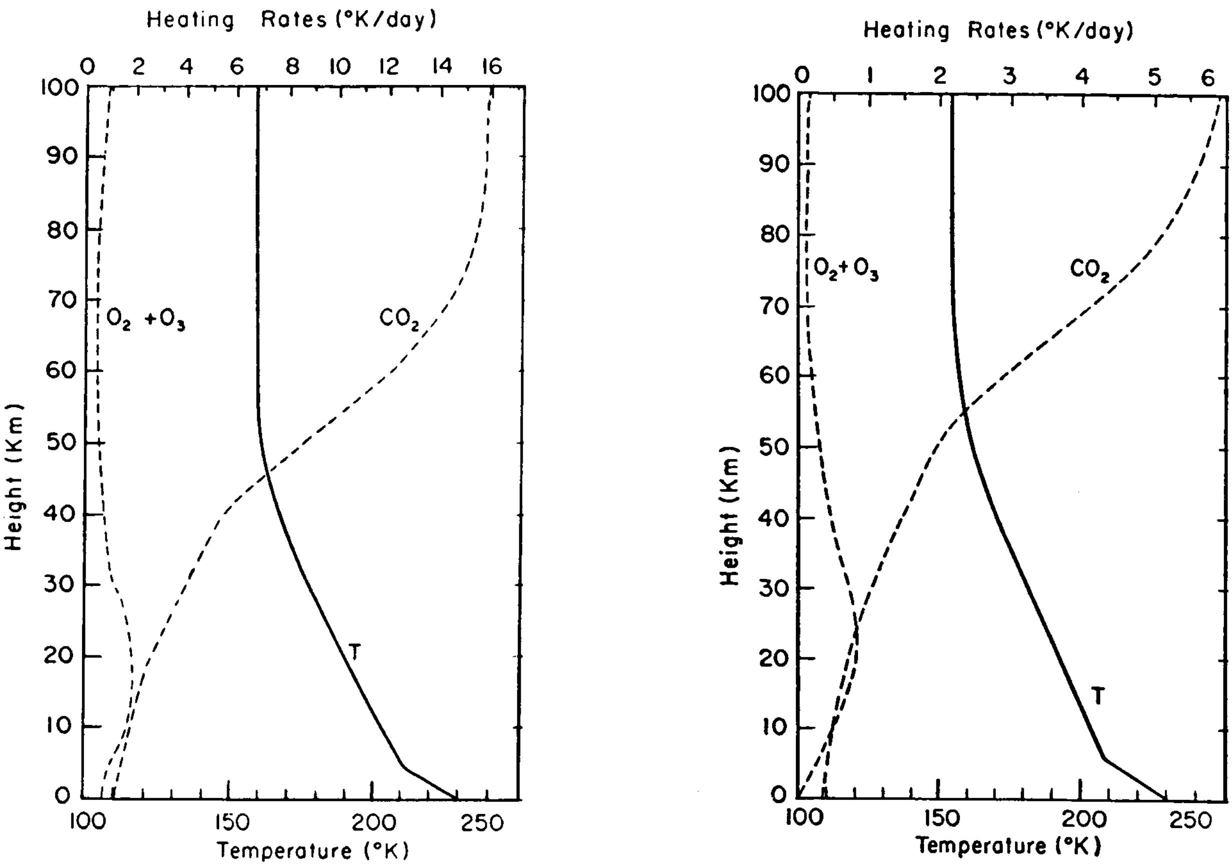


Fig. 1 -- Vertical temperature distribution on Mars as calculated by Goody and by Chamberlain compared with the vertical distribution in the Earth's atmosphere (from Chamberlain 1962).

of absorption of solar radiation by small amounts of oxygen and ozone. A number of different possible combinations of surface pressure, carbon dioxide, and oxygen concentration were tried, but the results were relatively insensitive to reasonable changes in these parameters. Figure 2 shows their results in two of these cases. Except for different assumed surface temperatures there is a close resemblance between these temperature profiles and Goody's.

The depth of the troposphere cannot be determined on the basis of radiative equilibrium alone, since convection on both small and large scales helps to determine the tropopause height. These calculations suggest, however, that a troposphere some 2 or 3 times as deep as the Earth's is likely. By terrestrial analogy, we may expect that this tropospheric layer behaves as a single dynamical system, in the sense that the whole region would act as a heat engine. In this heat engine, solar energy received near the ground in equatorial regions or in the summer hemisphere increases the internal and potential energy. The latter are converted to kinetic energy of the horizontal winds, and in the process, heat is transported upward and horizontally to heat-sink regions. A small portion of the kinetic energy may be transported upward out of the troposphere and be reconverted to internal plus potential energy. The potential plus internal energy produced in this refrigerator-like process would be destroyed by radiation. The corresponding combination of heat engines and refrigerators in the Earth's atmosphere has been discussed in detail by Newell (1965). We shall confine our attention to the tropospheric heat engine on Mars.



(a) surface pressure 10 mb, 44% CO_2, 0.4% O_2

(b) surface pressure 30 mb, 9% CO_2, 0.106% O_2

Fig. 2 -- Vertical temperature distributions for the lower atmosphere of Mars (from Prabhakara and Hogan 1965).

3. Diurnal heating and tides

One class of large scale motions which might be important on Mars are thermally driven tides. Observations (Sinton and Strong, 1960) indicate that the diurnal surface temperature oscillation is on the order of 100^{o}K. If this large amplitude is associated with a very large-amplitude diurnal component of vertical heat flux, significant tidal wind and pressure systems may be expected.

One way of estimating the diurnal component of the small scale vertical heat flux is to compare the observed surface temperature variations with those computed from a theory in which convective heat flux is taken into account. The amplitude and phase of the observed temperature wave can be matched by adjusting two parameters: the heat storage capacity of the ground and a heat exchange coefficient for the atmosphere (Leovy, 1965). A relatively simple theory that has been applied successfully to the Earth's atmosphere is that of Lonnqvist (1962, 1963). Lonnqvist assumes that heat exchange -- radiative, conductive, or convective -- between the ground and the atmosphere or space can be represented by Newton's law of cooling. Such a relation is valid for convective transfer under conditions of forced convection and steady winds. The heat balance condition at the ground (z = 0) takes the form

$$k\left(\frac{dT}{dt}\right)_{z=0} + \sigma T^4_{z=0} - R_b + h_c[T_{z=0} - T_a] = S \qquad , \qquad (1)$$

where the first term is heat flux into the ground; k is the soil's thermal conductivity to be determined. The second term is black-body emission from the ground. The third is back radiation from the atmosphere; the fourth is convective transfer between ground and atmosphere; h_c is a convective parameter to be determined, and T_a is a constant. S is the effective insolation. For an atmosphere without significant attenuation, the equation for S can be written

$$S = S_o(1 - A)(\sin\varphi \sin\delta + \cos\varphi \cos\delta \cos\omega t) \quad , \qquad (2)$$

where S_o is the solar constant at the distance of Mars, A the visible albedo, φ the latitude, δ the solar declination, ω the diurnal frequency, and t is time. Of course S vanishes when the right side of (2) becomes negative. The black-body emission law can be linearized with only small error. The shape and amplitude of the computed temperature wave near the equator then depends on only two parameters:

$$A = (\pi\,\beta)^{-1}(\cos\varphi \cos\delta) \cdot S_o(1 - A)$$

and

$$r = h/\beta \quad ,$$

where

$$\beta = (1/2\ \rho c\ k\omega)^{\frac{1}{2}} \quad ,$$

ρ and c are the density and specific heat capacity of the top few centimeters of soil, and h is the sum of h_c and a constant obtained from linearizing the black-body emission law.

Figure 3 shows two surface temperature curves obtained in this way which fit Sinton and Strong's observations fairly well. The value of β is so low that, for any reasonable values of density and specific heat capacity, k must be of the order of 10^{-4} cal/cm sec ^{o}K or less. This is so low that soil particle size in the top few centimeters must be comparable to the mean free path of air molecules -- a few microns at the probable surface pressure of Mars.

The parameter h_c can be estimated from the two attempted fits to the data shown in Fig. 3, and the corresponding most likely amplitude of the small scale convective heat flux is in the range 2--5 x 10^{-3} cal/cm^2 sec, or between 15 and 40 per cent of the peak insolation.

One can estimate the amplitude of thermal tides that this diurnally varying heat input would produce. A detailed study of possible Mars tides has been carried out by Craig (1964), who points out that the absence of any temperature maxima in the Martian stratosphere makes resonance amplification of the tides very unlikely. Calculation of tidal amplitudes should therefore not be very sensitive to the details of the vertical temperature structure. Craig computes the amplitude of the <u>diurnal</u> tide, which would arise from convective heating using a particularly simple model atmosphere and concludes that the maximum ratio of the diurnal tidal-pressure amplitude to the mean surface pressure should be about 1/600 or about the same as the corresponding quantity for the <u>semidiurnal</u> tide in the Earth's atmosphere. The convective heating assumed by Craig was only 1/3 to 1/8 as much as that estimated above, however. As long as the heating is confined to a thin layer near the surface -- a few kilometers deep, or less -- the exact distri-

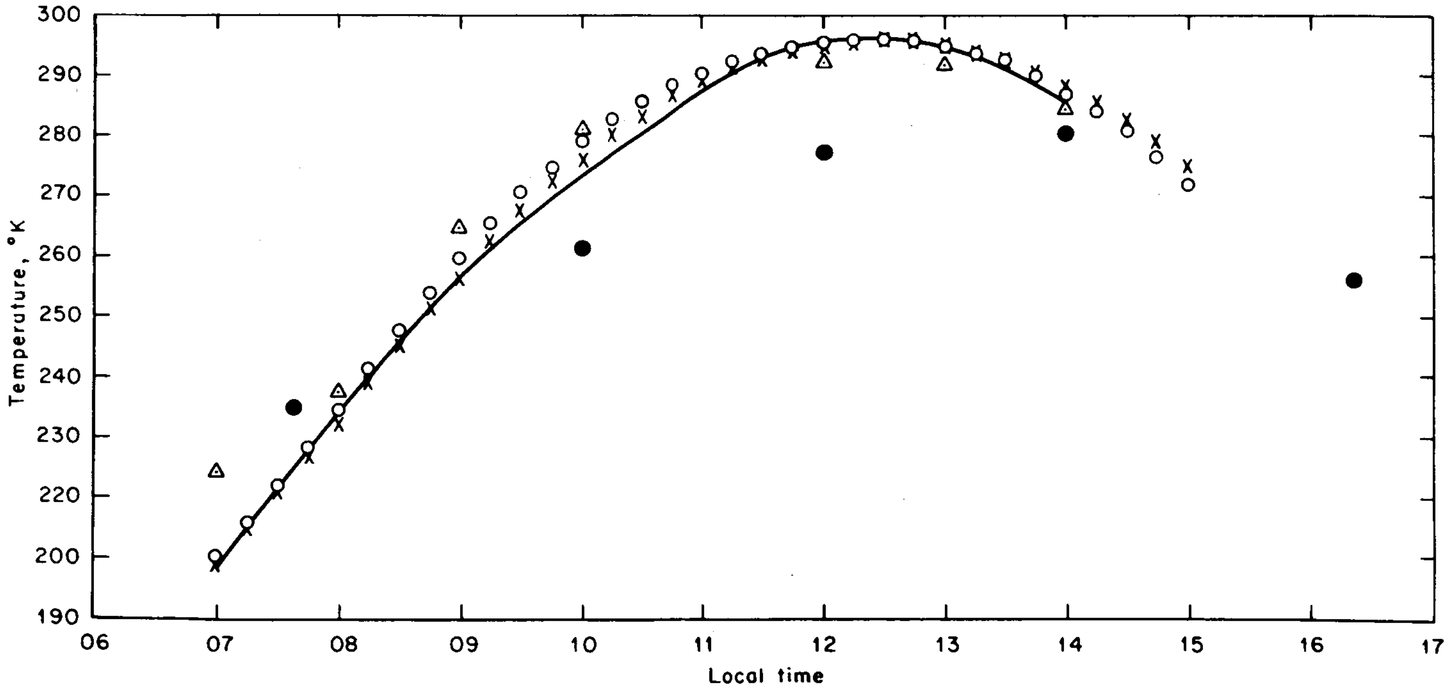


Fig. 3 -- Theoretical fit of Sinton and Strong's observations. Solid line is the average of their four best temperature curves; (x) are the fit for A = 236.5, r = 6, open circles are for A = 379, r = 10. The triangles are for data taken predominantly on dark areas. The solid circles represent the average of earlier observations by Coblentz and Lampland as presented by Gifford (1956).

bution of the heat input with height is unimportant; only its amplitude affects the tidal amplitude. It follows that the Martian diurnal tide may be as much as 8 times as large as the terrestrial semidiurnal tide, and could be associated with near surface winds of 3 to 4 meters per second.

The diurnal tide is mainly excited by heating in thin layers; the semidiurnal tide is excited by heating through deeper layers. The absorption of solar radiation by CO_2 thus contributes to the semidiurnal tide. Comparison of the solar heating rates calculated by Prabhakara and Hogan with the corresponding quantity for the Earth's atmosphere suggests that this component of tidal forcing is comparable on the two planets. Qualitatively then, one would expect that the semidiurnal tidal amplitude on Mars would be comparable to that on Earth. More detailed calculations are not justified at this time because of the lack of detailed knowledge about the vertical temperature distribution. It is worth noting that very large tidal winds (greater than about 10 m/s) appear to be ruled out by the lack of indication for such oscillations in the observations of cloud drifts.

4. Differential heating at the solstices

We come now to consideration of the atmospheric response to latitudinal and seasonal variations in heating. The results of a calculation by Mintz (1962) of the two most important heat balance components -- net incoming radiation and net outgoing radiation -- for the Martian southern hemisphere summer solstice are shown in Fig. 4. A calculation of this kind is more straightforward for Mars than for the Earth because of the virtually complete absence of clouds and atmospheric water vapor

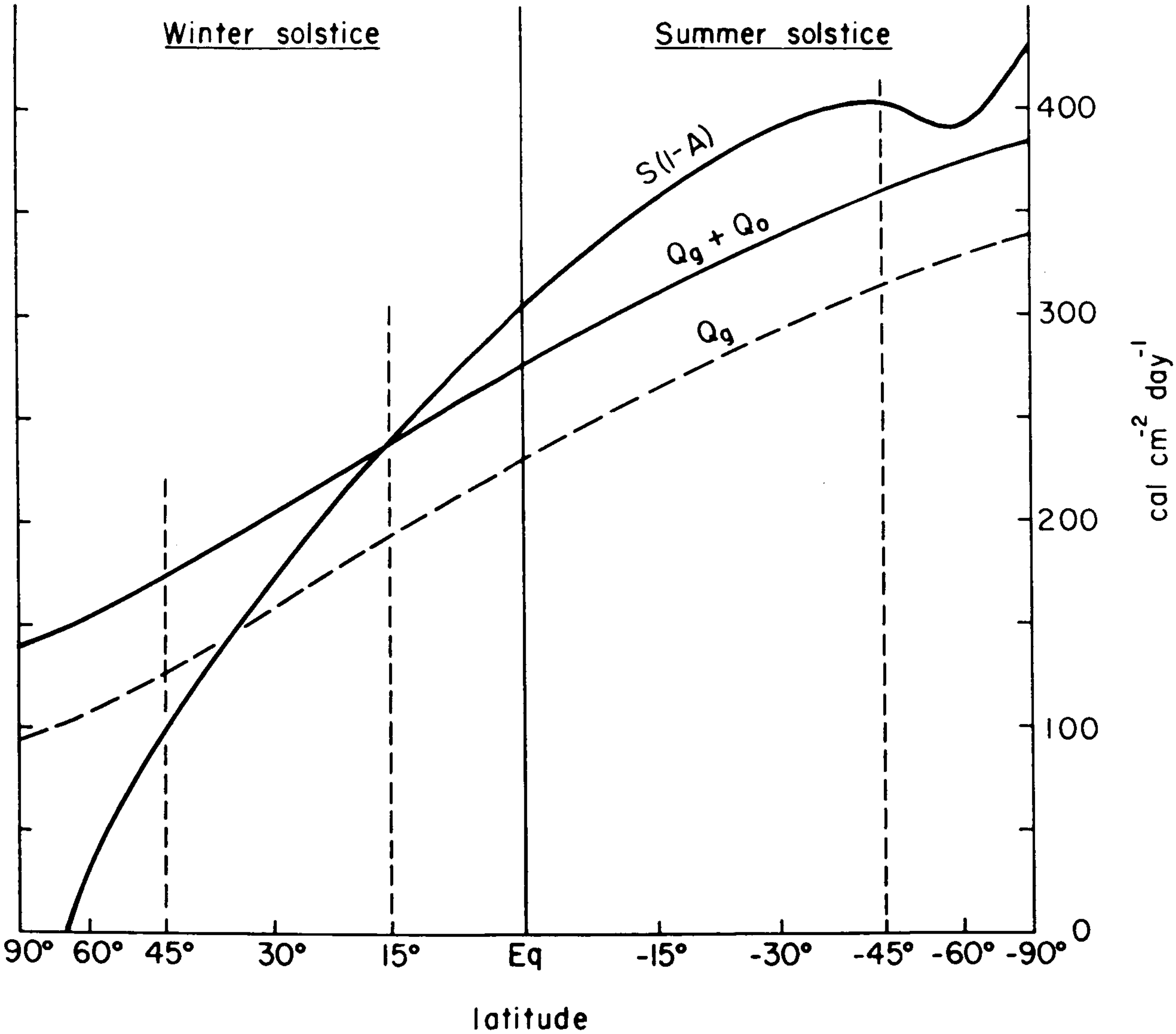


Fig. 4 -- Martian heat balance components. S(1 - A) is the net incoming radiation; Q_g is the radiation to space from the surface, $Q_g + Q_o$ is the total radiation to space from the surface and the atmosphere (from Mintz 1962).

on Mars. Furthermore, the difference between these two heat balance components, the net radiation excess, can be interpreted directly in terms of a heat transport requirement in the case of Mars. This is not possible for the Earth because of the large amounts of energy stored and transported by the oceans (seasonal storage of heat by the Martian soil must be negligible). Figure 5 shows the corresponding heat balance components for the northern hemisphere winter and summer of the Earth as computed by London (1957). It is ironic perhaps that no similar study for the southern hemisphere has been published yet, probably because of the uncertainties in cloud and water vapor distributions. Both planets show a net radiation excess in the summer and a deficit in the winter, but although such excesses and deficits can be interpreted as heat sources and sinks for the atmosphere on Mars, no such interpretation is possible for the Earth because of heat storage in the oceans.

Since the net radiation excess on Mars $Q(\varphi, t)$ can be assumed to be balanced by atmospheric transport, one can write:

$$Q(\varphi, t) = \frac{1}{a \cos \varphi} \frac{\partial}{\partial \varphi} \left\{ \cos \varphi \cdot 2\pi a^2 c_p P_o g^{-1} \cdot \overline{[v(T + gz\, c_p^{-1} + Lq\, c_p^{-1})]} \right\} , \qquad (3)$$

where φ is latitude, a is the planetary radius, P_o the surface pressure, and g is the acceleration of gravity. The bracketed quantity gives the advection of temperature T, potential energy gz, and latent heat energy Lq, where q is the specific humidity and L the latent heat of condensation. The advection is effected by the meridional wind component v.

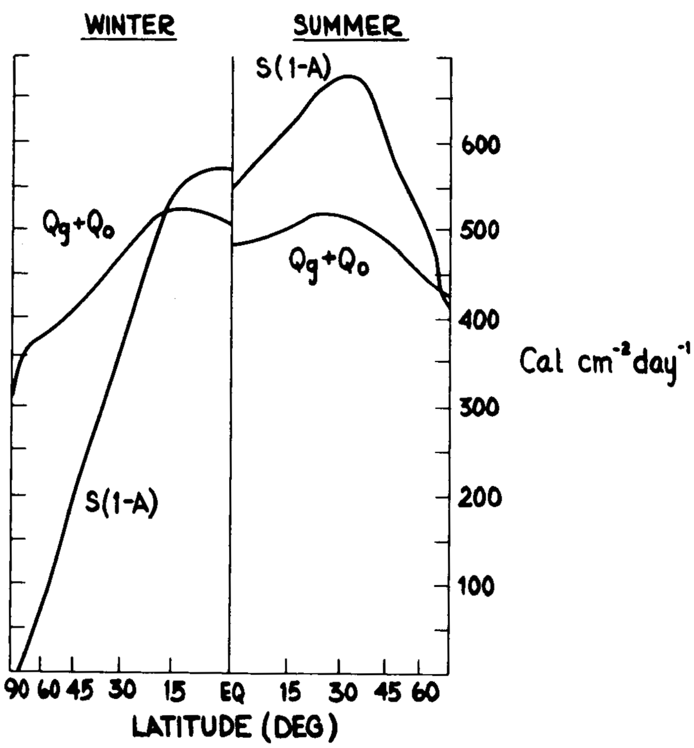


Fig. 5 -- London's data for the net incoming and net outgoing radiation on the Earth.

The bar indicates averaging over longitude, pressure, and time at a fixed season and latitude.

Noting that Lq can be neglected on Mars, this expression can be rewritten in the form,

$$H \equiv \left\{ \overline{v[T + gz\, c_p^{-1}]} \right\} = [\cos\varphi \cdot 2\pi a\, c_p\, P_o\, g^{-1}]^{-1} \cdot \int_{-\pi/2}^{Q} Q(\varphi,\, t)\, d(\sin\varphi) \quad . \qquad (4)$$

H is a normalized specific enthalpy transport requirement, which may be used to compare the intensities of circulations required on different planets. The right-hand side of Eq. 4 actually overestimates H for the Earth because of the oceanic heat storage. Nevertheless, it is instructive to compare the quantities,

$$[\cos\varphi \cdot 2\pi a\, c_p\, P_o\, g^{-1}]^{-1} \cdot \int_{-\pi/2}^{\varphi} Q(\varphi,\, t)\, d(\sin\varphi) \quad ,$$

for Mars and the Earth. This is done in Fig. 6 for an assumed Mars surface pressure of 30 mb. Evidently the solstice's specific enthalpy transport requirement, which is a measure of the intensity of the thermally driven large scale circulation, is more than twice as large on Mars as on the Earth. This conclusion is strengthened when the effects of oceanic transport and storage, and latent heat transport are taken into account; it is also strengthened if the actual surface pressure on Mars is less than 30 mb. The values of H deduced in this way will be used to estimate the probable magnitudes of the winds which provide the heat transport.

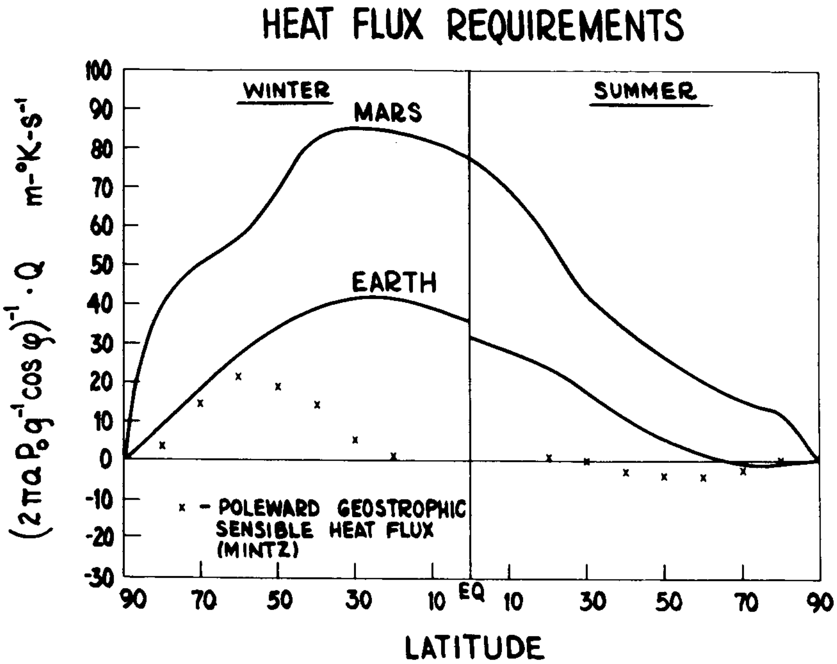


Fig. 6 -- The specific enthalpy flux requirements for the Earth and Mars. The poleward sensible heat flux calculated by Mintz (1954) and indicated in the diagram gives only the contribution from eddies. Any contribution from mean meridional circulations is not included in his data.

5. General circulation at the solstices

When the Rossby number and the ratio of horizontal scale to planetary radius are both small (R_o, $L/a \ll 1$), a considerable simplification of the hydrodynamical problem is possible, provided also that the parameter,

$$\epsilon = \frac{V^2}{gD\kappa} R_o^{-2} = \frac{f^2 L^2}{gD\kappa} \quad ,$$

is of order unity or less. (D is the characteristic depth of the circulation system.) The simplification is known as the quasi-geostrophic theory. (See, for example, Charney and Stern, 1962; Phillips, 1963; Pedlosky, 1964.) These conditions are likely to be satisfied at middle and high latitudes on Mars, although not so well satisfied as on the Earth -- nevertheless, we shall make use of this theory in a qualitative discussion of the Mars winds. The quasi-geostrophic theory replaces the complete system of hydrodynamic and thermodynamic equations with a single equation and a single physical principle -- the conservation of potential vorticity, q, defined by

$$q = \frac{\partial^2\psi}{\partial x^2} + \frac{\partial^2\psi}{\partial y^2} + (D/L)^2 \cdot \frac{1}{\bar{\rho}} \frac{\partial}{\partial z}\left(\bar{\rho}\, \epsilon \frac{\partial\psi}{\partial z}\right) \quad , \tag{5}$$

where ψ is the stream function for the geostrophic wind, and $\bar{\rho}$ is the horizontally averaged density. The conservation of potential vorticity is expressed by the relation,

$$\frac{\partial q}{\partial t} + \frac{\partial\psi}{\partial x}\frac{\partial q}{\partial y} - \frac{\partial\psi}{\partial y}\frac{\partial q}{\partial x} = 0 \quad , \tag{6}$$

where y is distance northward, x is distance eastward.

Charney and Stern have shown that, as a consequence of this equation, small disturbances in a basic zonal flow are necessarily stable if $\frac{\partial q}{\partial y}$ is of one sign and the latitudinal temperature gradient vanishes on the horizontal lower boundary. If the latitudinal gradient does not vanish, temperatures which fall toward the pole are destabilizing, and temperatures which rise toward the pole are stabilizing (Pedlosky, 1964). Because the gradient of f tends to dominate $\frac{\partial q}{\partial y}$, it is normally of one sign, and the effect of the temperature gradient at the ground seems to be the most important factor determining stability or instability. Results from other studies (Burger, 1962) suggest that small disturbances are always unstable if temperatures decrease toward the pole, the degree of instability increasing rapidly with the temperature gradient above a certain critical value of the latter. On the other hand, if temperatures increase toward the pole, small disturbances would fail to grow and in addition, any disturbances initiated at low levels by topography or local heating irregularities would be rapidly damped with height (Charney and Drazin, 1961). These theoretical differences between the two temperature gradient regimes are supported by experience in the terrestrial atmosphere. Thus we would expect that the summer hemisphere on Mars, in which daily mean temperatures increase toward the pole, would be stable and nearly zonally symmetric. The heat transport would be accomplished by a mean meridional circulation. On the other hand, the circulation at middle and high latitudes in the winter hemisphere should be dominated by large-scale quasi-horizontal eddies which transport the heat. Momentum-balance considerations would then require westerly zonal winds in the poleward portion of the winter hemisphere and near the surface in the summer hemisphere, and easterly winds elsewhere.

We may now estimate the order of magnitude of the winds required to satisfy the specific enthalpy transport requirement. In the region where large scale eddies predominate, we have approximately

$$H \sim \overline{vT} \approx \sigma(v)\ \sigma(T)\ \rho(v,\ T) \qquad , \tag{7}$$

where $\sigma(v)$ and $\sigma(T)$ are standard deviations of meridional wind and temperature at fixed latitude and season at some representative mid-tropospheric height. The correlation coefficient $\rho(v, T)$ may be large for these eddies, and by terrestrial analogy, a value of $\rho(v, T) \sim 0.3$ appears reasonable. According to the quasi-geostrophic theory, wind and temperature are related by the thermal wind equation, so that

$$\sigma(T) \sim \frac{fL}{R}\left(\frac{D^*}{D}\right)\sigma(v) \qquad , \tag{8}$$

where R is the gas constant and D^* is the scale height for Mars. Then

$$\sigma(v) \sim \left\{ \left[\frac{R}{fL\rho(v,\ T)}\right]\left(\frac{D}{D^*}\right) H \right\}^{\frac{1}{2}} \qquad . \tag{9}$$

Assuming that the correlation coefficient and the scale, which are determined by the mechanics of the instability process,‡ are roughly comparable for Mars and Earth and that $D^* \sim D$, for middle latitudes we find that the ratio of standard deviations is

‡Actually L will be a function of κ; see Mintz (1962).

$$\sigma_m(v)/\sigma_e(v) \sim [(RHf^{-1})_m/(RHf^{-1})_e]^{\frac{1}{2}}$$

$$\sim (H_m/H_e)^{\frac{1}{2}} \gtrsim 2 \qquad ,$$

where the subscripts m and e refer respectively to Mars and the Earth. The inequality is necessary since the values of H_e derived in Section 4 exceed the actual transport requirement. Equation 9 refers to the eddy velocity; the magnitude of the zonal wind velocity U is determined by a balance between loss of zonal momentum by vertical eddy stress, $K\,\frac{\partial U}{\partial z}$, where K is a vertical eddy stress coefficient, and production of zonal momentum by the large scale eddies. Assuming the kinematics of the unstable waves to be similar on the two planets, this production is proportional to $[\sigma(v)]^2$. Thus

$$\left[\left(K\frac{\partial U}{\partial z}\right)_m \bigg/ \left(K\frac{\partial U}{\partial z}\right)_e\right] \sim \left[\left(\frac{KU}{D^*}\right)_m \bigg/ \left(\frac{KU}{D^*}\right)_e\right]$$

$$\sim [\sigma_m(v)/\sigma_e(v)]^2 \qquad ,$$

and

$$(U_m/U_e) \sim [(KD^{*-1})_e/(KD^{*-1})_m] \cdot [H_m/H_e] \sim 6 \qquad , \qquad (10)$$

assuming $K_m \sim K_e$. Thus zonal winds of several hundred meters per second are possible at one-scale height above the Martian surface. The suface zonal wind ratio, on the other hand, would be comparable only to $(H_m/H_e)^{\frac{1}{2}}$.

For the mean meridional circulation regime of the winter hemisphere, it is somewhat more difficult to estimate velocities. It follows from Eq. 4 that

$$H \sim v(\gamma - \gamma_a)\ D_e \qquad , \qquad (11)$$

where v is the mean meridional wind component, $(\gamma - \gamma_a)$ is the difference between the actual and the adiabatic lapse rates (the adiabatic lapse rate is g/c_p), and D_e is the depth of the Ekman friction layer,

$$D_e \sim \pi(2Kf^{-1})^{\frac{1}{2}} \qquad (12)$$

(Taylor 1915). Near the equator D_e must be replaced with $D \sim D^*$. D_e and $(\gamma - \gamma_a)$ are difficult to estimate, but as an illustration of the magnitudes involved, we may take the reasonable values $H \sim 50$ m/s deg, $(\gamma - \gamma_a) \sim 2^{\circ}$/km, $D_e \sim 2$km, giving $V \sim 20$ m/s. This is a substantial mean meridional wind, and would be associated with a still larger zonal component.

Figure 7 illustrates these ideas schematically. The winter hemisphere contains strong zonal west winds that increase with height, except for a shallow belt of easterlies near the ground in low latitudes. The summer hemisphere is dominated by easterlies increasing with height -- except for a relatively shallow belt of suface westerlies. The meridional circulation associated with the eddy regime has descending motion equatorward of the west wind maximum and ascending motion poleward of the maximum. This circulation is a consequence of the production of zonal momentum by the eddies. In summer, a shallow poleward flow takes place beneath a deep, gradually ascending, return flow.

Although the Mariner 4 results (Kliore <u>et al.</u>, 1965) would alter some of the quantitative estimates given above through the surface

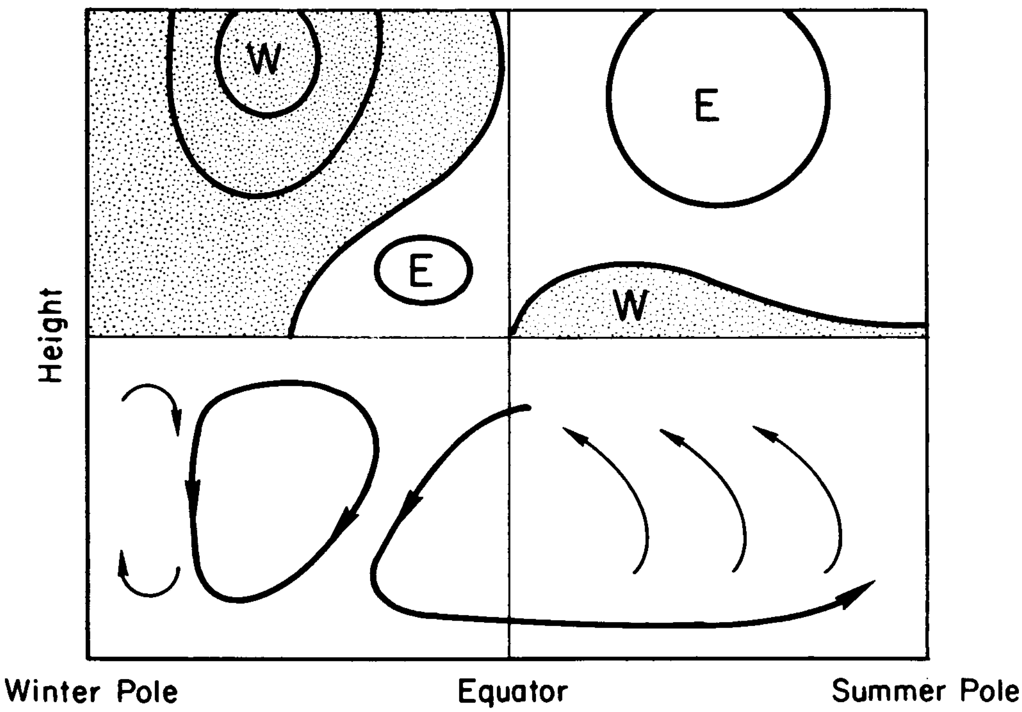


Fig. 7 -- Schematic diagram of solstice circulation on Mars. Top half is zonal circulation, bottom half is meridional circulation.

pressure that enters the expression for H_m, the main features shown in Fig. 7 still appear to be very likely.

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